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KINEMATICS, COMPOSITION AND THERMODYNAMICS OF THE MESOPAUSE, TURBOPAUSE REGION OF THE ATMOSPHERE (BETWEEN 80 AND 120 km) RELATED TO THE AEROASSISTED ORBIT TRANSFER VEHICLE (AOTV) OPERATIONS

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TABLE OF CONTENTS

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LIST OF ILLUSTRATIONS

TECHNICAL MEMORANDUM

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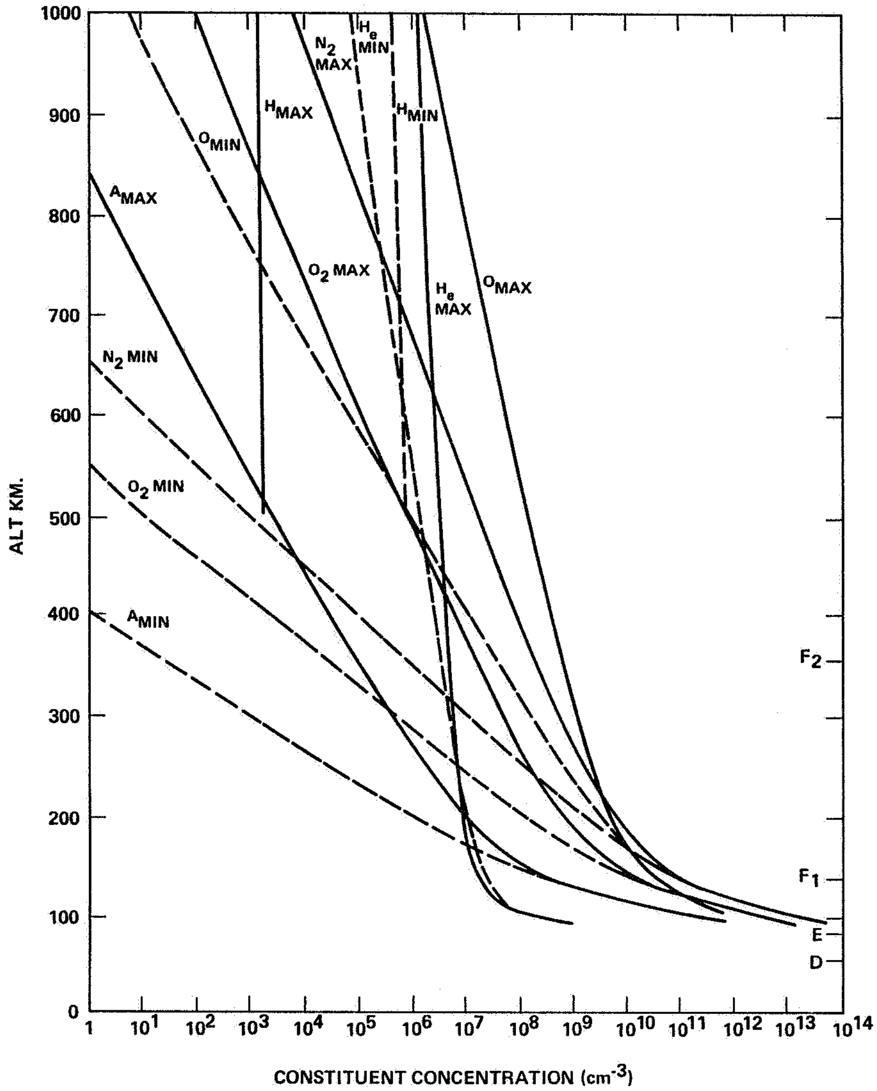
I. INTRODUCTION

The uncertainties in the atmospheric density variation in the mesopause-turbopause region (80 to 120 km), according to Walberg [1], lead to control system design problems for the AOTV relative to the amount of control authority required to deal with the unpredictable variation in density. The mesopause-turbopause interface region (80 to 120 km) of the atmosphere is frequently used as a boundary between the thermosphere and mesosphere for models of the atmosphere. Often these boundary conditions, according to Offermann, et al. [2], are assumed to be constant or almost constant. This assumption is not generally valid. In addition, the data base of neutral gas composition measurements is too small to permit such conclusions. There is general agreement that the major constituents, N_2 and O_2, are well mixed below the turbopause around 100 km, and that their mixing ratio is very near to its sea-level value. The location of the turbopause and its variations is not adequately understood. In this transition region, the number of experimental data sets of the total gas density and the constituents of smaller abundance is very limited. The instrumentation accuracy of measurements of the constituents, N_2, O_2, Ar, and CO_2, by mass spectrometers is believed to be no better than ±35 percent at all altitudes [2].

II. DISSOCIATION OF MOLECULAR OXYGEN AND MOLECULAR DIFFUSION

The processes of dissociation of molecular oxygen into atomic oxygen and of molecular diffusion, both occur within the height region between 80 and 120 km, also cause the modeling of this height region to be more complex than for other regions of the atmosphere. Consequently, the agreements and recommendations for a description of this height region in the 1976 U.S. Standard Atmosphere were generated only after considerable effort and revisions. The description is still open to question. The computed height distribution of molecular nitrogen, atomic and molecular oxygen, argon, helium and atomic hydrogen for the MSFC/J70 Model Atmosphere to heights up to 1000 km is depicted in Figure 1.

Barring long-term climatic changes, it appears unlikely that future data will suggest the need for significant changes in a mean static model atmosphere for 45 N. latitude at heights below 80 km. As indicated by Minzner [3], however, the probability that future data will support the 1976 U.S. Standard Atmosphere model is significantly decreased for heights between 100 and 120 km. Any future changes in a mean atmospheric model in this altitude range would, of course, create the need for changes in greater altitudes.



FOR: MIN SOLAR CONDITIONS: 0400 hrs. USING $F_{10.7}$ = 70 & Ap = 0
MAX SOLAR CONDITIONS: 1400 hrs. USING $F_{10.7}$ = 230 & Ap = 35
(MSFC–J70)

Figure 1. Constituent number density.

Results of measurements of atmospheric ion and neutral composition between 100 to 160 km were extensively reviewed by Danilov, et al. [4]. The measurements of atmospheric composition were made by a radio frequency mass spectrometer. The observations show the existence of strong diurnal variations of the atmospheric composition in the lower thermosphere and in the altitude range of the turbopause.

III. VARIATION OF TURBOPAUSE HEIGHTS AND NUMBER DENSITIES ASSOCIATED WITH TURBULENCE

Results of mass spectrometer measurements of the turbopause height (h_t) in the equatorial region were discussed by Danilov [5]. He found that this height varies from 92 to 115 km. No pronounced dependence of h_t on season, local time, or solar activity was determined. He noted that there appears to be a tendency for h_t to decrease with increasing temperature at 120 km.

The number of measurements of turbulence data is limited. However, it is well known that atmospheric turbulence plays an important role in the mesosphere and lower thermosphere, influencing the thermal balance of the upper atmosphere as well as the distribution of the different atmospheric constituents. Data from radio meteors, noble gas ratio analysis, and luminescent cloud analysis reveal that no definite conclusions or systematic variations of the turbulence regime can be made.

As mentioned earlier, the higher the temperature gradient between the mesopause and the bottom of the thermosphere the more "forbidden" is the development of turbulence at these altitudes, and so the turbopause height is lower. Thus, the vertical temperature gradient is one of the most important factors, which, together with the winds and convection, determine turbulence intensity in the upper mesosphere and lower thermosphere.

Figure 1 shows outputs of the MSFC Global Reference Atmosphere Model (GRAM)] for individual constituent number densities.

Figure 2 depicts the kinetic temperature as a function of geometric altitude as presented in the U.S. Standard Atmosphere [7].

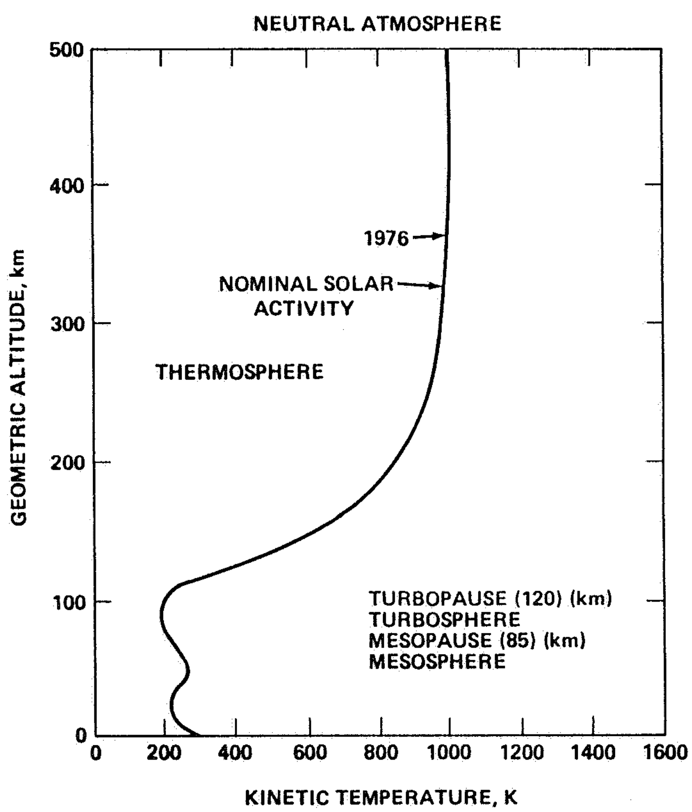


Figure 2. Kinetic temperature as a function of geometric altitude.

The turbopause is the boundary between complete mixing and complete diffusive separation and may be different for different constituents. Schuchardt, et al. [8], have shown that at high latitudes, the turbopause is lowest in winter conditions, thus reflecting the winter helium bulge. The turbopause varies greatly between about 90 and 115 km.

IV. STRUCTURE OF THE UPPER ATMOSPHERE

The variations and uncertainties of the structure of the upper atmosphere are so large and so complex that a discussion of the physical bases of the structure of the upper atmosphere is relevant. Brasseur and Solomon [6] and Van Zandt, et al. [10] provide the additional details on the physics of the upper atmosphere.

Brasseur and Solomon [6] describe the theoretical and thermal behavior of the atmosphere and indicate that the fundamental equations of fluid mechanics must be employed. As is well known, the circulation of the Earth atmosphere is governed by three basic principles: the Newton laws of motion and the conservation of energy and mass. The second law of Newton describes the response of a fluid to external forces. In a frame of reference which rotates with the Earth, it is given by:

$$\frac{d\vec{V}}{dt} + \frac{1}{\rho} \nabla p + 2\vec{\Omega} \times \vec{V} = \vec{g} + \vec{F} \quad , \tag{1}$$

where $\vec{V}$ represents the vector speed of an air parcel, p is pressure, ρ is the mass per unit volume, $\vec{\Omega}$ is the angular rotation of the Earth, $\vec{g}$ is acceleration of gravity, $\vec{F}$ is the frictional force due to viscosity, and t is time. The terms in equation (1) each correspond to a different force acting on the fluid. These are the pressure gradient force ($1/\rho\ \nabla p$), the Coriolis force ($2\vec{\Omega} \times \vec{V}$), the gravitational ($\vec{g}$) and the frictional ($\vec{F}$) forces (these conventionally represent the forces per unit mass in the meteorological studies). The second equation is the conservation of energy (first law of thermodynamics):

$$c_p \frac{dT}{dt} - \frac{1}{\rho} \frac{dp}{dt} = Q \quad , \tag{2}$$

where T is the temperature, c_p represents the specific heat at constant pressure, and Q is the net heating rate per unit mass. The third fundamental equation is conservation of mass (also called continuity):

$$\frac{d\rho}{dt} + \rho \nabla \cdot \vec{V} = 0 \quad . \tag{3}$$

V. WINDS AND TURBULENCE

In the determination of mesospheric winds, the complexity of the chemical and photo-chemical processes is another factor that makes it extremely difficult to ascertain wind flow characteristics with any acceptable accuracy. All theoretical wind models which compute the seasonal wind structure rely on the separation of density (ρ_1) and velocity (V_1) where 1 refers to different periods from one day to a year [9]. Uncertainties in the knowledge of the wind velocities lead to much greater uncertainties in the horizontal divergence of the wind which must be known for an evaluation of the vertical diffusion velocities. All of these difficulties reduce the reliability of existing three-dimensional models in describing the effects.

The value of the eddy diffusion coefficient in the turbulent regions of the lower thermosphere and altitude of the upper limit of the turbulence, called the turbopause, each play an important rôle in the calculation of minor constituents in the thermosphere [11]. In effect the altitude at which diffusion equilibrium starts to be established depends on two factors. If, in the turbulent regions, the eddy diffusion coefficient (which depends on the energy of the turbulence) is everywhere greater than the molecular diffusion (which depends on altitude and molecular weight), this equilibrium can only be established in the turbopause. If, on the other hand, the eddy diffusion coefficient becomes lower than the molecular diffusion coefficient of the minor constituent, within a given turbulent region, diffusion equilibrium will start being established at that point. The value of the eddy diffusion coefficient in the lower thermosphere has been deduced theoretically from a study of the heat balance. The theoretical deduction and empirical estimate lead to a value in the neighborhood of 10^7 cm^2/s.

In this study, the eddy diffusion limit of sodium can be considered as an indicator of the turbopause or of the altitude above which the eddy diffusion coefficient varies substantially during the night.

In order to achieve this, Teitelbaum and Blamont [11] reveal the existence of two highly different wind regimes during the night between 95 and 110 km altitude, especially from the standpoint of the variation in mean kinetic energy with altitude. The relations between the variation in kinetic energy with altitude, the energy dissipated, and the eddy diffusion coefficient were established.

Figure 3 shows an example of the variation in r.m.s. wind velocity with altitude between 89 and 115 km for Period I (from dusk to 10 p.m.) and Period II (from 10 p.m. to dawn). In making these calculations, 24 wind hodographs were employed, 12 for each period. The vertical velocities were ignored. It can be seen that the mean kinetic energy is nearly always lower in Period I than in Period II in the altitude range. Moreover, it appears to decrease more rapidly with altitude during Period I.

Teitelbaum and Blamont [11] presented measurements of radio wave absorption in the ionosphere which were used to estimate the coefficients of horizontal and vertical turbulence diffusion at about 100 km altitude. For central Europe, values of 9×10^6 cm^2 sec^{-1} were obtained for the vertical and 7×10^9 cm^2 sec^{-1} for the horizontal directions.

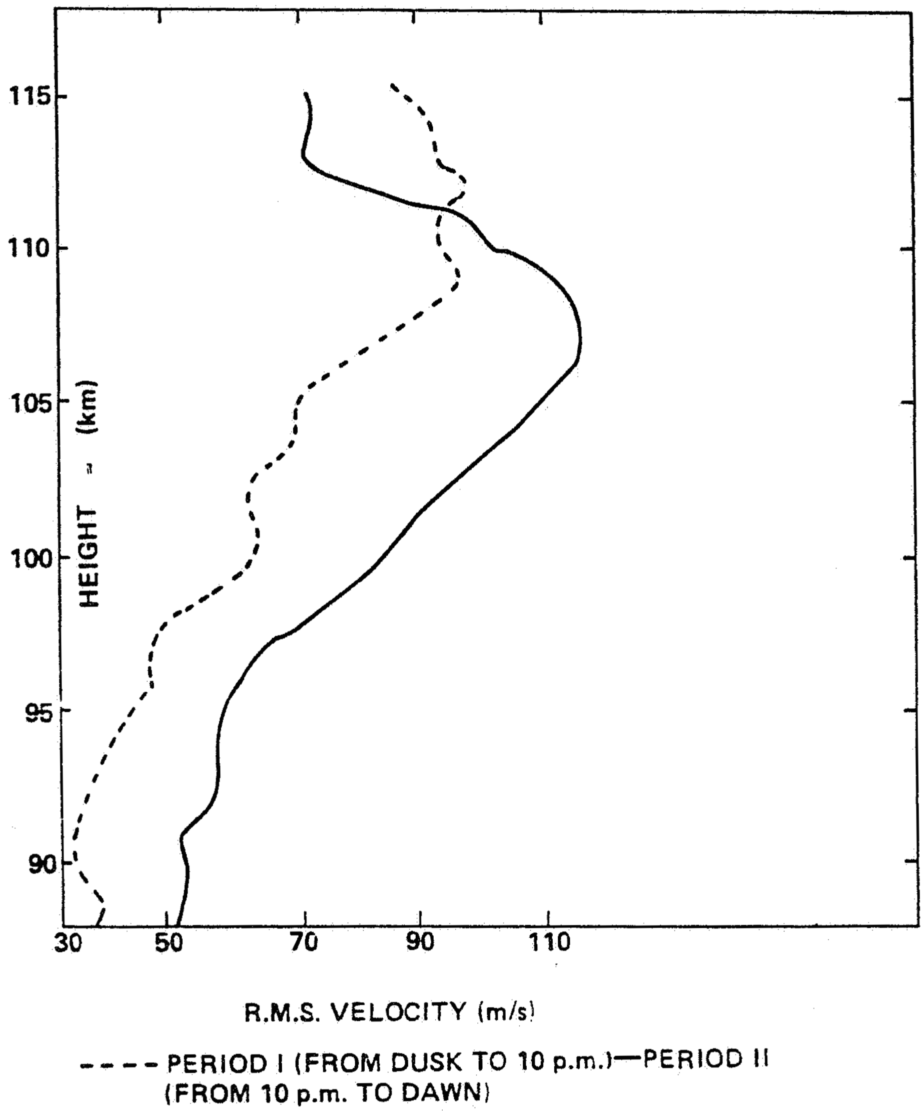


Figure 3. R.M.S. wind velocity versus altitude for wind measured over Yuma, Arizona.

The radar echoes in the mesosphere arise from two separate processes [12,13]. The first involves the action of small-scale atmospheric turbulence operating on an existing vertical gradient of electron density. Turbulence is thought to arise from the breakdown of upward propagating gravity waves and tides. These "turbulence" echoes occur primarily during daytime since the requisite electron density is strongly reduced at night. The second echoing process makes use of short-lived ionization trails deposited by meteoric particles.

Below the mesopause (∿ 85 km) turbulent mixing keeps the composition of the atmosphere homogeneous, except for traces like water vapor and ozone, which are affected by processes which are fast compared with turbulent mixing. The percentages of the number densities of the major species at sea level are given in Table 1.

TABLE 1. PERCENTAGES OF THE NUMBER DENSITIES OF MAJOR ATMOSPHERIC SPECIES AT SEA LEVEL

N_2	O_2	A^{40}	Sum of Species	$\bar{M}$ (Mean Molar Weight)
78.08	20.95	0.93	99.96	28.96

Above the mesopause molecular oxygen is dissociated by solar ultraviolet radiation with wavelengths shorter than 1759 Å. This process is so rapid that turbulence cannot keep O_2 and the resulting O mixed. Therefore, the number density of atomic oxygen increases rapidly with height to a peak at about 100 km. The percentage concentration of atomic oxygen also increases, but instead of reaching a peak it continues to increase with height until, above about 150 km, the atmosphere is dominated by atomic oxygen.

As long as there is no tendency for the constituents to be separated by diffusion, the ratio of the number of oxygen atoms (whether free or bound) to the number of nitrogen atoms (whether free or bound) is unchanged by dissociation. The condition is expressed by the relation

$$\frac{[O] + 2[O_2]}{2[N_2]} = \frac{20.95}{78.08} = 0.2683 \quad . \tag{4}$$

The brackets denote number densities. It it assumed here that molecular nitrogen is not appreciably dissociated in the atmosphere.

Although above the mesopause, turbulence is not sufficient to keep O_2 and O mixed, it still mixes the constituents faster than diffusion separates them. The rate of diffusion increases exponentially with height and proportionally with the mean free path, whereas the rate of mixing by turbulence appears to decrease sharply near 100 km. Therefore, above the level called the "turbopause," each constituent obeys a separate barometric equation with its own "partial scale height," $H_i = RT/M_i g$; that is,

$$\frac{p_i(h)}{p_i(h_t)} = \exp\left[-\int_{h_t}^{h} \frac{dh}{H_i(h)}\right] = e^{-z_i(h,h_t)} \quad , \tag{5}$$

where i denotes the ith constituent, h_t is the height of the turbopause, and $z_i(h,h_t)$ is the reduced height of the ith constituent; z_i is, of course, proportional to M_i.

VI. GRAVITY WAVES AND THEIR COMPLEX DEVELOPMENT AND BEHAVIOR PLUS IMPLICATION FOR AOTV OPERATION CONTROL IN THE 70 TO 90 km ALTITUDE REGION

It is now generally accepted, as discussed by Walterscheid [14], that the dynamics and structure of the mesosphere and lower thermosphere are strongly influenced by the propagation of internal gravity waves. Lindzen [15] and Matsuno [16] have suggested that breaking gravity waves in the mesosphere could be the physical mechanism for deceleration of the zonal wind system above the stratopause.

The heat input by gravity waves is a dominant term in the energy balance equation. The internal gravity waves produce a variation in total density. As Walberg [1] indicated, this may lead to control system design problems for the AOTV relative to the amount of aerodynamic control authority required in the typical orbit transfer vehicle maneuver from High Earth Orbit (HEO) and Low Earth Orbit (LEO).

The basic idea underlying the saturation theory presented by Lindzen [15] is that gravity waves grow exponentially with height until the perturbation lapse rate becomes sufficiently large to cancel the mean (stable) lapse rates. At this time, convective instabilities occur that act to dissipate the wave motion and cancel the mean (stable) lapse rate and future wave growth. The induced wave dissipation results in a vertical divergence of the gravity wave momentum flux, causing an acceleration of the medium toward the phase velocity of the wave.

In general, then saturation refers to any mechanism that acts to limit or reduce wave amplitudes due to their growth with height. For simplicity, however, it is assumed that mesospheric saturation is a convection instability and has a narrow spectrum in nature.

Schoeberl [17] reported qualitatively agreement with Walterscheid [14] on results of the dissipation of a gravity wave with 1000-km zonal and meridional wavelengths. The deceleration is substantially ~ 45 m s^{-1} deg^{-1} at 97 km; and a peak cooling rate of ~ 3.5 K day^{-1} is produced in the 95 to 100 km zone with heating below 95 km ~ 1.2 K day^{-1}. Thus, the dynamical coupling of the troposphere is extended to the lower thermosphere.

Fritts, et al. [18] presented an analysis of gravity waves with a period of 10 hr. They found the wave to be circularly polarized with vertical and horizontal wavelengths ~ 14.5 and 1590 km, and a phase velocity of 44 m sec^{-1} to the South and Reynolds stress ~ 1.77 m^2 sec^{-2}, resulting in a mean flow deceleration of $\bar{V}_t \sim 25$ msec^{-1} day^{-1}. Based both on observed amplitude and direction of propagation of the 10-hr wave and on simultaneous observation of large-amplitude, high-frequency vertical motions, it was concluded that gravity waves must be contributing in a significant way to gravity wave saturation and zonal flow deceleration.

From the continuous Fourier spectrum, Battaner [19] identified internal gravity waves. The emission of the 557.7 nm emission line of atomic oxygen is mainly produced at 90 to 100 km, where internal gravity waves are especially important. This is approximately the saturation level (below 90 km), where the amplitude growth is effected by nonlinear aspects. The heat input by gravity waves is a dominant term in the energy balance equation. Also, gravity waves, either directly or through a mechanism of turbulence generation, enhance the mixing ability of the atmosphere in this region where molecular diffusion is a competitive effect.

Internal gravity waves produce variations in the density as well as in the concentration of atomic oxygen. Hence, variation in the green line airflow are to be expected, which means that observation of this emission feature would be an additional method of studying internal gravity waves.

VII. GROUND-BASED REMOTE SENSORS, ROCKET-BORNE SENSORS AND SPACE-BORNE SENSORS MEASURE DENSITY IN THE 80 TO 120 km REGION

Density in the mesosphere can indeed change drastically within a short period of time [20]. Philbrick [20] used Robin spheres to measure density. The Robin is a passive/reflective 1 m inflatable sphere that is ejected from a rocket nose cone at 120 km apogee and radar tracks its fall. Useable atmospheric data between 90 and 35 km altitude is produced based on drag tables. Philbrick [20] indicated that changes in atmospheric density by as much as 20 percent may occur in a period of a few hours. These changes appear to occur in layers of a few kilometers in vertical thickness.

The Rayleigh-Lidar sounder, according to Chanin et al. [21], has been developed to probe the middle atmosphere. The system has demonstrated its ability to provide density and temperature profiles in the 30 to 90 km region. It has been limited to ground-based operations, but it could be used in the future from a space-borne platform. The altitude range fo the Lidar sounding is fixed by the upper level of the aerosol layer. The upper limit is given by the signal-to-noise ratio, thus explaining the great care taken to reduce the sky-background which is the main contribution to noise. After a succession of improvements through the years, the range was pushed up from 70 km in 1981 to 90 km and occasionally to 95 km (not operational). With a still increased sensitivity, one may be able in the future to detect the eventual presence of aerosols of meteoritic origin around 90 to 100 km, but, as for now, their abundance has never been measured by Lidar.

Hedin et al. [22] indicated that the MSIS-83, Mass Spectrometer and Incoherent Scattered data, is a revision of the MSIS-79. The model is based on temperature, density, and composition data from a comprehensive summary of rocket flights, seven satellites, and five incoherent scatter radars. The model extends the previous description of neutral parameters to the base of the thermosphere in a continuous manner. As the altitude decreases, composition approaches lower atmosphere values while yearly, and to a lesser extent, daily variations in temperature and density are in reasonable agreement with previous results for the lower atmosphere.

Composition data gathered by rocket mass spectrometer and EUV absorption measurements were taken from the summary by Offerman [23] and augmented by Ackerman et al. [24], Cooley and Reber [25], and Trinks et al. [26]. Neutral composition measurements were obtained from the satellites ESRO 4, AEROS A, AEROS B, and Atmosphere Explorer C. A comprehensive summary of total density and (molecular) temperature data from rocket-borne pressure gauge, grenade, and falling sphere measurements which reach up to the lower thermosphere have been reported by Minzner et al. [27] and Theon [28].

In this transition region, as previously pointed out, the number of experimental data sets of the total gas density and the constituents of smaller abundance is very limited.

VIII. DISCUSSION OF AEROASSISTED ORBIT TRANSFER VEHICLES (AOTV) AND UNCERTAINTY OF DENSITY

The AOTV uses aerodynamic forces to produce an orbital change with an expenditure of energy significantly smaller than that associated with an extra-atmospheric propulsive maneuver. According to Walberg [1], almost any mission that involves changes in orbital altitude and/or inclination in the vicinity of an atmosphere-bearing planet is a candidate for aeroassist. The typical orbit transfer vehicle (OTV) where the vehicle is initially in HEO – geosynchronous, undergoes a sizeable velocity decrement to LEO. The velocity decrement is more efficiently achieved through a combination of aerodynamic drag and propulsion than through propulsion alone.

One of the central issues, as outlined by Walberg [1], is the amount of aerodynamic control authority in the guidance and control system design required to deal with the unpredictable variations in atmospheric density. The magnitude of the atmospheric dispersions that must be dealt with is indicated in Figure 4. Only present estimates of the mean and range of density over an 11 year period are shown and are expressed as a relative difference (percent) from the 1962 Standard Atmosphere. The mean density constitutes a systematic variation that is predictable for a given month and solar activity by the NASA/MSFC Global Reference Atmosphere Model (GRAM) (1980). The range constitutes a random variable that is not predictable. The geomagnetic storm variation is for an extreme event that occurs once or twice per cycle and is not predictable even a few days in advance. It usually occurs during periods of medium or high solar activity and the effects do not propagate into the lower more dense atmosphere. The range of density below 100 km does not depend on solar activity. The unpredicatable density variation, such as those shown

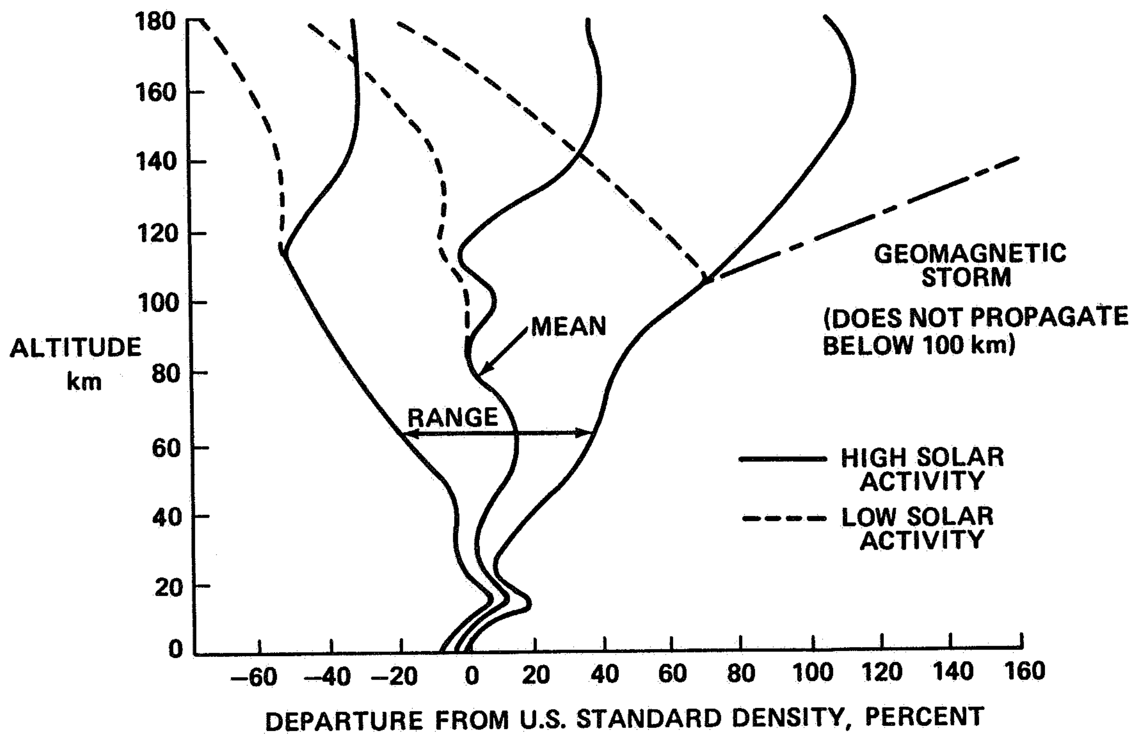


Figure 4. Atmospheric density dispersions based on the 1962 U.S. Standard Atmosphere, Summer.

in Figure 4, is one of the prime concerns for the AOTV design. In addition, data from Shuttle flights 2, 4, and 6 have further shown the existence of so-called "potholes in the sky" as illustrated in Figure 5. The NASA/MSFC Global Reference Atmosphere Model (GRAM) (1980) contains the provision for model outputs of such profiles to use in AOTV design tradeoff and sensitivity analysis.

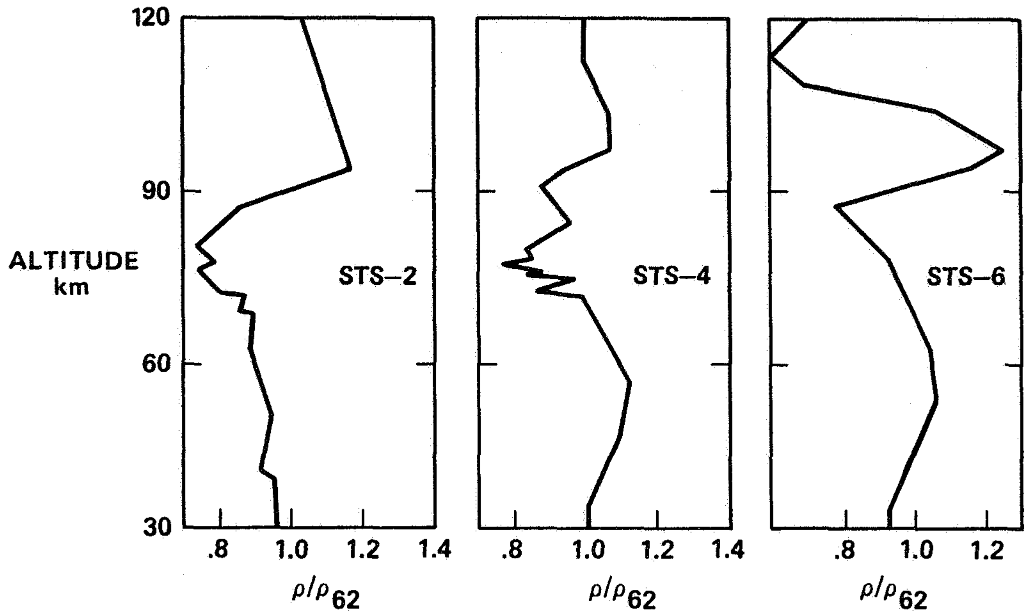


Figure 5. Shuttle-derived densities compared to the 1962 U.S. Standard Atmosphere.

There are large density excursions, which occur over relatively narrow altitude ranges, that will pose a significant challenge to the AOTV design flight control systems. The question is how much drag variation or lift/drag capability is required to compensate for these density variations. According to Walberg [1], present indications are that fairly sophisticated adaptive control laws may be required for AOTV. Given the influence of random gravity waves on this region of the atmosphere the ability of adaptive control systems to respond in sufficient time to be effective may be the deciding factor. It remains to be seen whether such control laws can be implemented into an operational vehicle and still accommodate the range of unpredictable density profiles which may be encountered.

IX. CONCLUDING REMARKS

The mesopause-turbopause (80 to 120 km) region of the atmosphere is frequently used as a boundary between the thermosphere and the mesosphere for models of the atmosphere. In this transition region between continuum flow below about 90 km and free molecular flow above this altitude, the number of experimental data sets of the total gas density and the constituents of smaller abundances are very limited. The initialization of these data bases in this transition region is important since uncertainties may lead to significant errors in the computation of the atmospheric total density at greater altitudes.

Experiments by Offermann [2], Minzner [3], Danilov [5], Strickland [29], and Jacchia [30] have indicated the turbopause height (h_t) may vary from 92 to 115 km and no dependence of h_t on season, local time, or solar activity is determined. Teitelbaum and Blamont [11] and Balsley and Riddle [12] have indicated the important role of turbulence in the mesosphere and lower atmosphere influencing the thermal balance of the upper atmosphere as well as the distribution of different atmospheric constituents. Data from radio meteors, noble gas analysis, and luminescent cloud analysis reveal no systematic variation of the turbulence regime can be made. The heat input by gravity waves is a dominant term in the energy balance equation through a mechanism of turbulence generation which enhances the mixing ability of the atmosphere. The internal gravity waves produce variations in the density as well as concentration of atomic oxygen. The uncertainty in the density variation according to Walberg [1] may lead to critical control systems design problems relative to the amount of aerodynamic control authority required to deal with the unpredictable variations in the density of the atmosphere.

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APPROVAL

KINEMATICS, COMPOSITION AND THERMODYNAMICS OF THE MESOPAUSE, TURBOPAUSE REGION OF THE ATMOSPHERE (BETWEEN 80 AND 120 km) RELATED TO THE AEROASSISTED ORBIT TRANSFER VEHICLE (AOTV) OPERATIONS

By Michael Susko

The information in this report has been reviewed for technical content. Review of any information concerning Department of Defense or nuclear energy activities or programs has been made by the MSFC Security Classification Officer. This report, in its entirety, has been determined to be unclassified.

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16. ABSTRACT

This report addresses the mesopause-turbopause region (80 to 120 km) of the atmosphere which is frequently used as a boundary between the thermosphere and mesosphere for models of the atmosphere. The initialization of models is important since uncertainties may lead to significant changes in the computations of total density at greater altitudes. In this transition region, the experimental data base for the total gas density and the constituents of smaller abundance is very limited. The turbopause height (h_t) may vary from 90 to 120 km and no pronounced dependence of h_t on season, local time, or solar activity is determined.

The importance of atmospheric turbulence is discussed and its important role in the mesosphere and lower atmosphere by influencing the thermal balance of the upper atmosphere, as well as the distribution of different atmospheric constituents, is presented. The number of measurements of turbulence at these altitudes is small. Data from radio meteors, noble gas ratio analysis, and luminescent cloud analysis reveal no definite conclusion or systematic variation of the turbulence region. The heat input by gravity waves is a dominant term in the energy balance equation. Also gravity waves, either directly or through a mechanism of turbulence generation, enhance the mixing ability of the atmosphere. Internal gravity waves produce variations in the density as well as concentration of atomic oxygen.

The uncertainty in the atmospheric density variation, according to Walberg (1985), leads to control system design problems for the AOTV relative to the amount of control authority required to deal with the unpredictable variation in density.

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